

Lagrangian Diagnostics of Tropical Deep Convection and Its Effect upon Upper-Tropospheric Humidity

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ABSTRACT

This study combines geostationary water vapor imagery with optical cloud property retrievals and microwave sea surface observations in order to investigate, in a Lagrangian framework, (i) the importance of cirrus anvil sublimation on tropical upper-tropospheric humidity and (ii) the sea surface temperature dependence of deep convective development. Although an Eulerian analysis shows a strong spatial correlation of ~ 0.8 between monthly mean cirrus ice water path and upper-tropospheric humidity, the Lagrangian analysis indicates no causal link between these quantities. The maximum upper-tropospheric humidity occurs ~ 5 h after peak convection, closely synchronized with the maximum cirrus ice water path, and lagging behind it by no more than 1.0 h. Considering that the characteristic e -folding decay time of cirrus ice water is determined to be ~ 4 h, this short time lag does not allow for significant sublimative moistening. Furthermore, a tendency analysis reveals that cirrus decay and growth, in terms of both cloud cover and integrated ice content, is accompanied by the drying and moistening of the upper troposphere, respectively, a result opposite that expected if cirrus ice were a primary water vapor source. In addition, it is found that an $\sim 2^\circ\text{C}$ rise in sea surface temperature results in a measurable increase in the frequency, spatial extent, and water content of deep convective cores. The larger storms over warmer oceans are also associated with slightly larger anvils than their counterparts over colder oceans; however, **anvil area per unit cumulus area, that is, cirrus detrainment efficiency, decreases as SST increases.**

1. Introduction

Water vapor is arguably the most important greenhouse gas; therefore, a thorough understanding of its global distribution is essential for making credible climate change projections. Despite its small concentrations, upper-tropospheric water vapor is particularly important in determining climate sensitivity to increasing CO_2 due to its radiative effects (Shine and Sinha 1991). The dependence of the upper-tropospheric vapor budget on tropical deep convection (DC) has been recognized at least since the study of Gray et al. (1975). Their finding that humidity in the tropics is generally high under convective conditions has later been confirmed by satellite studies. For example, Soden and Fu (1995) have found a strong positive relationship between the frequency of tropical deep convection and upper-tropospheric humidity (UTH) over a wide range

of spatial and temporal scales. Udelhofen and Hartmann (1995) have also demonstrated that UTH increases linearly with the amount of tropical upper-level clouds, but rapidly decreases with distance from cloud edge. More recently, Su et al. (2006) have observed strong positive correlations between annual means of UTH and tropical cirrus ice water content.

The existence of such apparently strong spatial relationships has led some investigators to conclude that the evaporation–sublimation of cloud condensate, particularly of cirrus ice, plays a significant role in moistening the tropical upper troposphere (Soden 1998; Su et al. 2006). Other diagnostic studies, however, have argued that the amount of cirrus ice is too small for substantial sublimative moistening (Luo and Rossow 2004; John and Soden 2006); hence, the upper-tropospheric vapor budget is, instead, more likely to be influenced by dynamical mechanisms responsible for the formation and maintenance of cirrus anvils (CAs; Soden 2004).

Modeling studies are also divided over the relative importance of various mechanisms influencing the tropical vapor budget. Sun and Lindzen (1993) have

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developed a conceptual model for tropical convection, concluding that the evaporation of hydrometeors pumped into the upper troposphere by deep convective towers is the chief moisture source for the large-scale subsiding flow. They have specifically ruled out detraining of saturated air as a significant process in maintaining the observed humidity distribution. In the radiative-convective model of Rennó et al. (1994), the equilibrium climate also strongly depends on the amount of cloud condensate available for evaporative moistening, which in turn is determined by the precipitation efficiency of clouds. Finally, a recent modeling study by Folkins and Martin (2005) indicates that the dominant source of water vapor to most of the tropical troposphere is evaporative moistening, which only diminishes in importance at elevations close to the tropopause.

Other investigators dispute the above findings. For example, Sherwood (1999) has concluded that upper-tropospheric moistening by cloud sublimation is small, proposing instead the redistribution of vapor from other parts of the atmosphere toward cloudy regions by a “pumping” mechanism arising from the local radiative effects of cirrus clouds. This interpretation lends support to results obtained from simulations and trajectory analysis that tropical UTH distributions can be explained simply on the basis of advection without inclusion of hydrometeor evaporation and ice-liquid water transport (Salathé and Hartmann 1997; Dessler and Sherwood 2000). Consequently, the concept of large-scale control has been introduced, which postulates that the subtropical and tropical moisture fields are governed primarily by large-scale processes, such as the three-dimensional wind field (Pierrehumbert and Roca 1998).

The relationship between tropical deep convection and UTH is further complicated by potential cloud feedback mechanisms accompanying a predicted increase in sea surface temperature (SST). The two most controversial studies both suggest a negative cloud feedback, although through significantly different processes. The “thermostat hypothesis” proposes an increase in the amount of deep convective cirrus anvils with SST (Ramanathan and Collins 1991). Reflective cirrus clouds reduce the solar radiation reaching the surface, thereby regulating the surface temperature. Conversely, the “iris hypothesis” speculates that increasing SST is accompanied by a decrease in detrained cirrus clouds resulting from enhanced precipitation efficiency in cumulus towers (Lindzen et al. 2001). The diminishing cirrus cover then allows more outgoing longwave radiation to escape into space, thus providing a negative feedback to regulate SST.

The outcome of the debate about processes and feedbacks required to determine tropical humidity distributions has far-reaching consequences. If a detailed representation of cloud microphysics is necessary in global climate models (GCMs), then climate change estimates from current models employing simplified moist physics might be highly inaccurate. However, if mechanisms such as hydrometeor evaporation and the thermostat-iris effect were proven to be of secondary importance, existing climate predictions could be accepted with higher confidence.

Motivated by the controversies outlined above, this paper evaluates the life cycle of tropical convective systems from a Lagrangian perspective by combining geostationary satellite imagery with cirrus cloud property retrievals and high-resolution microwave SSTs. Specifically, the detraining of cirrus ice water in tropical deep convection, the impact of its subsequent sublimation on UTH, and the sensitivity of convective life cycle to SST are investigated.

2. Data

a. Brightness temperatures and upper-tropospheric humidity

This study uses 6.7- μm water vapor ($T_{6.7}$) and 11- μm window channel (T_{11}) brightness temperatures from the geostationary Multifunctional Transport Satellite 1R (MTSAT-1R) from January 2006. Data have been provided by the Atmospheric Radiation Measurement (ARM) program, which archives hourly observations at 5-km resolution in the tropical western Pacific (TWP) domain (40°S–30°N, 80°E–160°W).

The water vapor channel is used for humidity retrievals and trajectory calculations, while the window channel is used for segmenting cloudy pixels into DC and CA. The actual quantity retrieved from clear-sky water vapor brightness temperature is the weighting function-averaged relative humidity corresponding to a deep, cloud-free layer roughly between 500 and 200 mb. Hereafter this parameter is given with respect to ice, and is referred to as UTH. Based upon simplified radiative transfer calculations, Soden and Bretherton (1993) derived the following analytic expression to relate clear-sky $T_{6.7}$ to UTH:

$$\ln\left(\frac{\text{UTH}p_0}{\cos\theta}\right) = a + bT_{6.7}, \quad (1)$$

where θ is the satellite view zenith angle, p_0 is a climatological base pressure, and $a = 34.35$ and $b = -0.126$ are empirical coefficients derived specifically for the MTSAT-1R water vapor channel. In comparison to Ra-

man lidar retrievals, UTH computed by (1) has a moist bias and rms difference both less than 10% (Soden et al. 2004). Moreover, as shown by Soden (1998), the spatial and temporal variations of UTH, which are the focus of this paper, are very robust and insensitive to small errors in the accuracy of the UTH retrieval.

b. Cloud properties

The corresponding cloud optical properties have been obtained from the Visible Infrared Solar–Infrared Split Window Technique (VISST; Minnis et al. 1995). Here, version 1 of the pixel-level, 8-km-resolution cloud ice water path (IWP), cloud-top pressure, cloud phase, and cloud mask products are used over the VISST TWP domain (20°S–10°N, 120°–180°E).

The accuracy of these optical retrievals, particularly of ice water path, is hard to establish. Analysis of microwave and optical cloud water path retrievals by Horváth and Davies (2007) has found large biases and rms differences, and weak correlations between the two techniques for mixed-phase, precipitating clouds. A comparison between VISST and a satellite–surface–microwave hybrid algorithm has shown that VISST IWP is biased high by $\sim 30\%$ for ice-over-water cloud systems (Huang et al. 2005). More recently, Dong et al. (2008) have compared VISST with ARM measurements in overcast stratus conditions and concluded that VISST can provide accurate water path retrievals during daytime; however, its skill is generally degraded during nighttime because split-window signals are effectively lost for thicker clouds.

In our analysis night retrievals do appear noisier; nevertheless, they yield the same mean results as day retrievals, hence, we only present combined day-plus-night results in section 4. In sum, absolute values of VISST IWP retrievals should not be blindly trusted; however, they should still convey useful information about the relative time evolution of ice condensate.

c. Microwave optimally interpolated sea surface temperatures

Our global sea surface temperature dataset has been provided by Remote Sensing Systems (RSS). This particular product is based on microwave observations whose through-cloud capability make them superior to traditional SST data derived from infrared measurements in cloud-free regions only. Here, we use version 2 of the daily, 0.25° (~ 25 km)-resolution optimally interpolated SST product blending measurements from Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) and Advanced Microwave Scanning Radiometer (AMSR) Earth Observing System

(EOS) (AMSR-E). The RSS retrieval technique is a multistage linear regression algorithm that considers the effect of diurnal warming and “normalizes” all SSTs to a daily minimum value, defined to occur at approximately 0800 LT (Wentz and Meissner 2000; Gentemann et al. 2003). The estimated rms accuracy of the retrieved SSTs is 0.5 K.

3. Methodology

To study the life cycle of tropical deep convection, a synergistic Lagrangian database has been created in which the frame of reference moves with the developing cloud systems. This is achieved by tracking water vapor features in consecutive $6.7\text{-}\mu\text{m}$ satellite images and compositing the resulting trajectories with the various datasets discussed in section 2. The end product is a large set of synchronized cloud trajectories collected over a 1-month period, describing the time evolution of upper-tropospheric humidity, cirrus ice water path, sea surface temperature, and anvil and deep convective cloud cover. The relevant aspects of the methodology are provided below in some detail.

a. Lagrangian tracking

Starting from an initial image at time $t = \tau$, the procedure selects “target” boxes the size of a typical GCM grid cell (51×51 pixels or $\sim 255 \times 255 \text{ km}^2$) and tracks them on the next ($t = \tau + 1 \text{ h}$) or the previous ($t = \tau - 1 \text{ h}$) image. Only sufficiently large cloud patterns with at least 10% deep convective cloud cover (defined in section 3b) are considered. For each target box a “destination” box yielding the highest spatial cross-correlation and passing a self-symmetry test is then determined. Finally, using the destination box as the new target box, the same procedure is repeated to construct 24-h trajectories forward and backward in time. To avoid multiple instances of the same cloud trajectory in our dataset, targets are selected from images at 0000, 0600, 1200, and 1800 UTC only.

The water vapor radiance patterns being tracked are generally confined to a narrow range of altitudes (approximately 200–350 mb) corresponding to the upper-tropospheric layer flow, but the exact height and thickness of the layer depend upon the nature of the target (Schmetz et al. 1995). This study focuses on patterns containing significant cloud cover, which tend to follow the motion of the cloud top.

b. Subdivision into deep convection and cirrus anvil

In each tracking image, pixels labeled as ice-topped clouds by VISST are classified as either a DC core or

CA following the methodology of Fu et al. (1990). Specifically, the following three distinct cloud clusters are identified in a two-dimensional histogram of T_{11} brightness temperatures and local noontime visible reflectance: (i) a cold and bright one corresponding mainly to deep convective cores, (ii) a moderately cold, dimmer, and more diffuse one corresponding typically to cirrus anvil clouds, and (iii) a warm and dark one comprising mostly of small, low-level clouds. Respective T_{11} and reflectance thresholds of 215 and 260 K, and 0.6 and 0.4 isolate these clusters quite well.

To sample the full range of diurnal variability, the final classification employs T_{11} measurements alone, because they are available both day and night. Therefore, we define deep convective cloud cover (DCC) and cirrus anvil cloud cover (CAC) as the percentage of pixels in a tracking box with $T_{11} < 215$ K and $215 \text{ K} < T_{11} < 260$ K, respectively. This brightness temperature-only cloud classification still separates cloud clusters fairly well and independent of SST despite a slight increase/decrease in cloud-top height/temperature over warmer oceans. It is found that our cloud classification results in a monthly mean DC amount of 6.2% in the TWP domain for January 2006, which is in good agreement with the 20-yr January mean International Satellite Cloud Climatology Project (ISCCP) DC amount of $6.3\% \pm 1.1\%$.

It is emphasized that because of the continuous variation of brightness temperature, there is no ideal threshold to perfectly separate DC and CA clouds. Fortunately, the distribution statistics of deep convective systems are relatively insensitive to the precise value of the T_{11} threshold within a reasonable range (Mapes and Houze 1993). It is also noted that split-window-corrected effective temperatures would more accurately identify the thinnest cirrus clouds than uncorrected brightness temperatures. However, the error in CA area introduced by semitransparency effects is typically less than 5%; therefore, the missed thin cirrus contributes little to the overall water budget (Machado and Rossow 1993).

A more important limitation is that in a fixed-sized domain more DCC necessarily implies less area available for CAC. In other words, as deep convection fills the tracking box its cirrus anvil increasingly falls outside the box, leading to a spurious negative correlation between instantaneous DCC and CAC as noted by Del Genio et al. (2005). Simply expanding the domain would not eliminate this artifact because it may mistakenly include high clouds that are not associated with the targeted convection. Instead, we perform *region-growing* calculations that determine only those cirrus clouds in the full image that have a direct path of con-

nectivity to the originally targeted DC, similarly to Machado et al. (1998). Given a starting location and a range of values for which to search, the algorithm finds all cells within the T_{11} array that are within the specified range and have some path of connectivity through these cells to the starting location. (Cloud cells are considered adjacent and connected when they share either a common edge or corner.)

Region-growing calculations are crucial when investigating the SST dependence of cirrus detrainment efficiency because this requires determining the absolute areas of CA and DC segments of arbitrary sizes and shapes. In section 4, plots labeled the “*region-growing method*” represent such computations with results expressed as cloud area (km^2) rather than cloud fraction. In contrast, the effect of ice sublimation on UTH can be best evaluated with gridded data; hence, we keep the fixed-box analysis in this case.

c. Tracking-box-mean IWP and UTH

The tracking-box-mean cirrus anvil IWP and UTH are calculated for each time step within a trajectory. Note that the whole-box average IWP, rather than the in-cloud average IWP, is the proper measure of the total condensate mass available for sublimation in each tracking box. The mean UTH is estimated strictly from clear pixels and only for domains with less than 95% cloud cover, due to strong cloud attenuation effects in the water vapor channel. Pixels are classified as cloud contaminated when $(T_{11} - T_{6.7}) < 25$ K (Soden 1998, 2004).

d. Identification and synchronization of individual deep convective events

Although each trajectory is started from a particular deep convective target, transitions from one cloud to another are possible with cross-correlation tracking. Therefore, every trajectory has to be tested for multiple deep convective events by analyzing the time evolution of DCC as illustrated in Fig. 1a by the solid line. Here, the cross marks the DCC in the initial target box at time $t = 0$, while $t > 0$ and $t < 0$ correspond to the forward and backward trajectory segments, respectively. As shown, four distinct DC events can be identified in this particular trajectory, represented by the black, red, green, and blue segments. Such DC events are then synchronized by resetting the time of their DCC_{max} to $t = 0$.

Individual DC events can be easily identified in the CAC evolution as well (dashed line). Interestingly, for a given cloud system the maximum CAC lags behind the maximum DCC by a few hours. This then finally brings us to introduce some terminology that

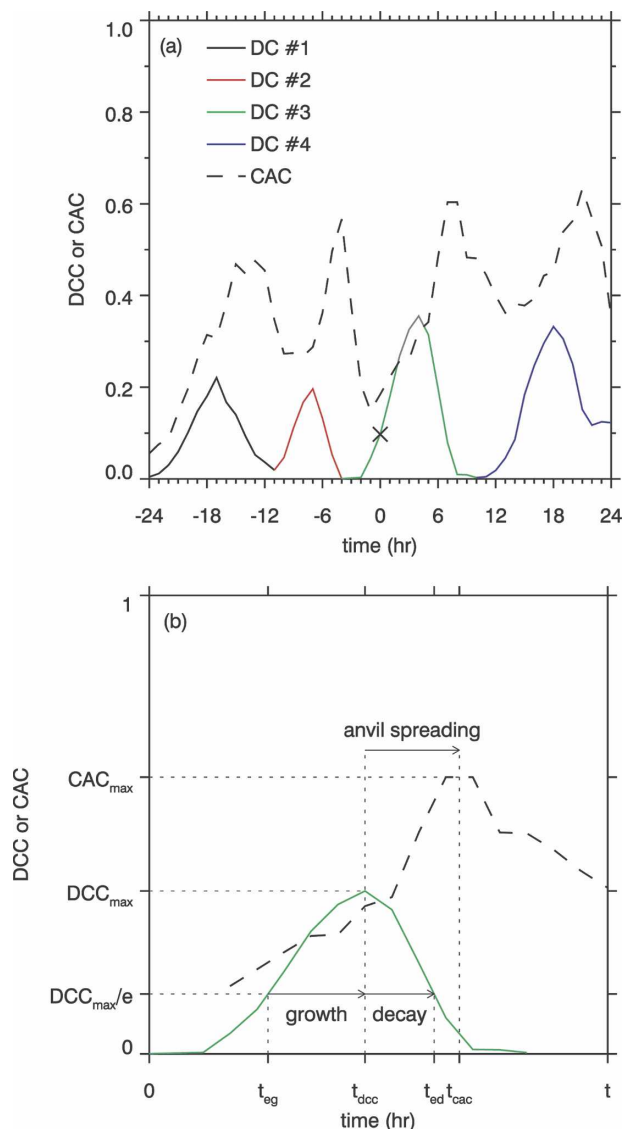


FIG. 1. (a) Evolution of DCC and CAC for a particular trajectory over a 48-h period. The cross marks the DCC in the initial target box while the colored segments correspond to the identified individual DC events. (b) Definition of time scales. Anvil-spreading time $t_{\text{cac}} - t_{\text{dec}}$: time difference between DCC and CAC peaks (similarly for IWP). Growth, $t_{\text{dec}} - t_{\text{eg}}$, and decay, $t_{\text{ed}} - t_{\text{dec}}$, e -folding times: time interval in which quantity increases or decreases by a factor of e . Lifetime: sum of growth and decay e -folding times.

will be used in the rest of this paper (see Fig. 1b). We define *anvil-spreading time* with respect to cirrus cloud fraction as the time difference between CAC_{max} and DCC_{max} or with respect to ice amount as the time difference between IWP_{max} and DCC_{max} . In addition, we characterize the evolution of DCC, CAC, and IWP with growth and decay e -folding times, the sum of which is defined as *lifetime*.

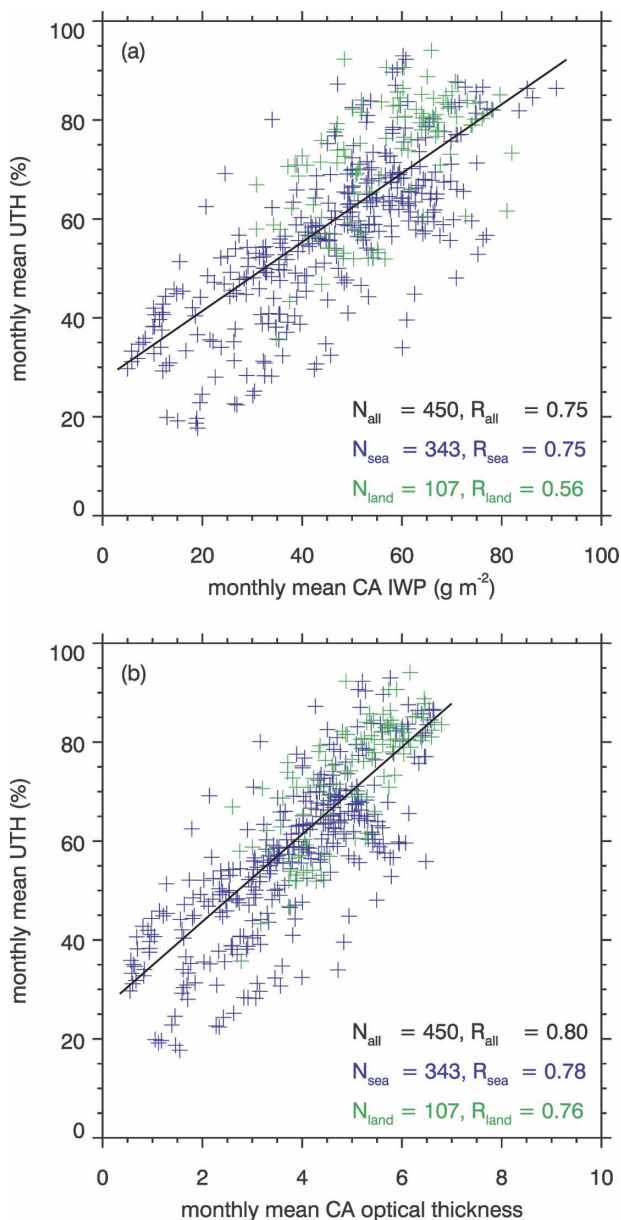


FIG. 2. Scatterplot of UTH vs (a) CA IWP, and (b) CA optical thickness for $2^\circ \times 2^\circ$ monthly means in the TWP domain. Blue and green plus signs correspond to ocean and land boxes, respectively, while the black line represents a linear fit to the entire dataset. Sample sizes (N) and correlations (R) are also indicated.

4. Results

a. Spatial correlation between UTH and anvil ice

Before the Lagrangian analysis, we present Eulerian spatial correlations between UTH and CA IWP, and UTH and CA optical thickness in Fig. 2 for $2^\circ \times 2^\circ$ monthly means. Ocean and land grid boxes are plotted in blue and green, respectively, while the black line

represents a linear fit to the entire dataset. Note that CA IWPs and optical thicknesses over land are at the higher end of the distribution and have a narrower range compared to oceanic values. (The same is true for DC ice content as well.) This finding is consistent with Zipser et al. (2006), who observed the strongest storms occurring preferentially over land.

Clearly, UTH shows a strong positive correlation with both measures of anvil Ci amount. When UTH is plotted against ice water path (Fig. 2a), the correlation is significantly lower over land ($R_{\text{land}} = 0.56$) than over ocean ($R_{\text{sea}} = 0.75$); however, when it is plotted against cloud optical thickness (Fig. 2b), the discrepancy between land and ocean correlations disappears. Because ice water path is proportional to the product of cloud optical thickness and ice crystal effective radius, these results point to potentially larger retrieval uncertainties in ice crystal size over land. This finding is consistent with Dong et al. (2008), who also observed reduced correlations between ground-based and VISST effective radius estimates at the ARM SGP site. Thus, the looser relationship between UTH and CA IWP over land is likely an artifact rather than a real signal.

As previously discussed, similarly strong correlations between deep convection and upper-tropospheric water vapor have been found in other observational studies as well. In addition, global climate models are also capable of reproducing this spatial relationship as shown by John and Soden (2006). Such apparently strong correlations have led some authors to conclude that sublimation of detrained ice can be a significant source of UTH (Soden 1998; Su et al. 2006). However, in the remainder of this paper we will attempt to show that while deep convection and local UTH are indeed connected, sublimative moistening does *not* play an important role in modulating the upper-tropospheric vapor budget, at least when averages in a deep layer between 500 and 200 mb are considered.

b. Mean trajectories for January 2006

The Lagrangian methodology presented in section 3 yields $\sim 14\,000$ trajectories, and $\sim 30\,000$ individual DC events for January 2006, resulting in ~ 2 DC cloud systems per every 48-h track, on average. To demonstrate our tracking method's ability to accurately characterize the upper-tropospheric circulation, Fig. 3a shows monthly mean trajectories binned in $4^\circ \times 4^\circ$ latitude-longitude boxes. (Here, monthly mean tracks are preferred to individual trajectories simply for clarity of presentation.) As a comparison, corresponding monthly mean upper-tropospheric winds from the National Centers for Environmental Prediction–National

Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) are presented in Fig. 3b on a $5^\circ \times 5^\circ$ grid. There is remarkably good qualitative agreement between the satellite tracks and model analysis in the depiction of the large-scale upper-level flow. North of the equator the strong easterlies of the trade winds dominate. South of the equator the picture is more complex. There is a divergence feature at 15° – 20°S , 135° – 140°E , west of which the flow is predominantly easterly and east of which it is mostly westerly. An upper-level anticyclone, with its outflow crossing the equator, is also evident at 10°S , 165° – 170°E . This semipermanent anticyclone is a well-known feature of the western Pacific tropical circulation in January (Sadler 1975).

The above monthly mean trajectories only depict the large-scale circulation pattern in a qualitative manner. For a quantitative assessment of our tracking method the reader is referred to Soden (1998), where a comparison with an operational “wind” algorithm designed as input to numerical weather prediction found the respective water vapor motion vectors well correlated ($r > 0.85$) with average absolute differences of $< 4 \text{ m s}^{-1}$ and biases of $< 1 \text{ m s}^{-1}$ in both east–west and north–south vector components.

c. Mean evolution of deep convective systems

The Lagrangian evolution of DCC, CAC, UTH, DC IWP, and CA IWP averaged over all trajectories and DC events is plotted in Fig. 4, with Figs. 4a, b showing fixed-box results while Fig. 4c corresponds to the region-growing algorithm. In this and subsequent figures the x axis is synchronized time with $t = 0$ referring to peak convection and the vertical bars depict the standard errors of the means.

As shown, DCC exponentially increases (decreases) before (after) the central time and is fairly symmetrical about its peak value. The evolution of CAC, on the other hand, is distinctly skewed, with its growth being slower than its decay. In Fig. 4a the local CAC minimum at $t = 0$ is the fixed-domain artifact discussed in section 3b, which disappears when the region-growing method is employed in Fig. 4c. Note that the maximum CAC occurs 4–5 h after peak convection, regardless of whether the fixed-box or the region-growing analysis is used. This lag characterizes the average anvil-spreading time.

In comparison to the Lagrangian study of Luo and Rossow (2004), the evolution of CAC in our dataset is similar to that of their cirrostratus category, but is significantly faster than that of their cirrus class. This might be due to our uncorrected brightness tempera-

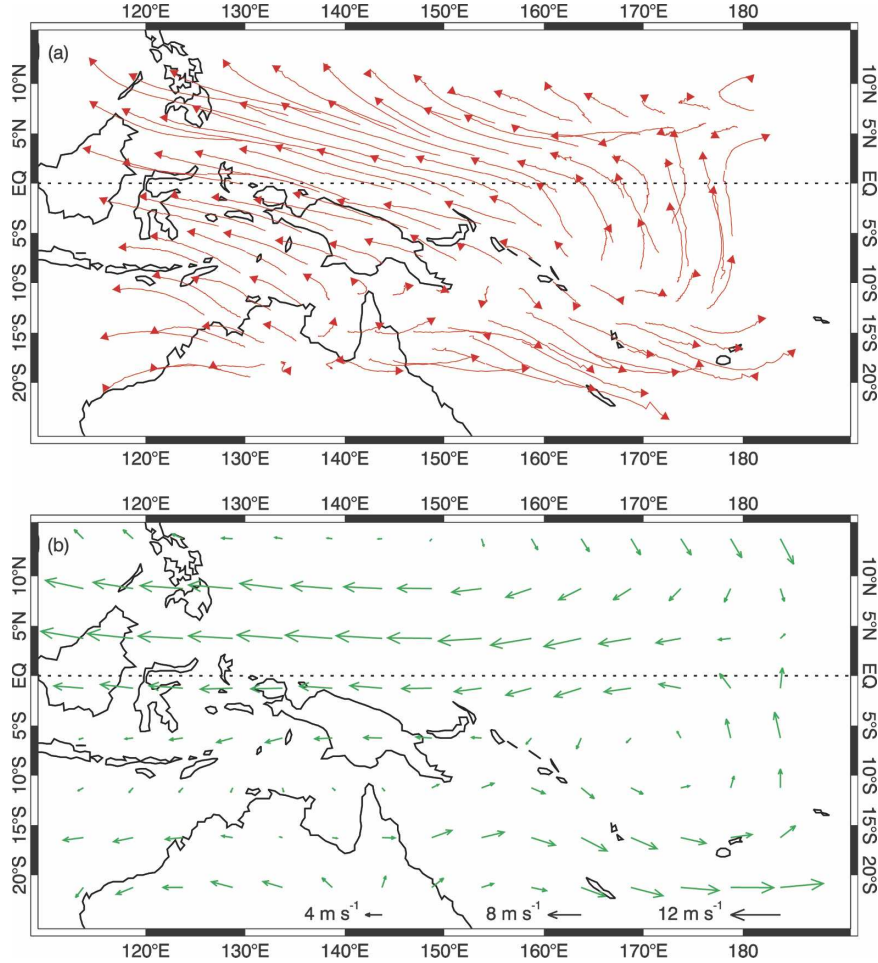


FIG. 3. Mean upper-level circulation in the TWP domain for January 2006: (a) average trajectories on a $4^\circ \times 4^\circ$ grid, and (b) average NCEP–NCAR reanalysis winds on a $5^\circ \times 5^\circ$ grid.

tures identifying only thicker parts of the anvil and missing the thinnest semitransparent cirrus. However, it is equally plausible that the discrepancy is caused by the relatively crude “tracking” method of Luo and Rossow (2004), who were unable to track the life cycle of any specific cirrus cloud; instead, they calculated forward air trajectories from 6-h-resolution reanalysis winds.

The evolution of DC and CA IWP closely resembles that of DCC and CAC (Fig. 4b), but with their relative magnitudes reversed: the maximum DC IWP is ~ 1.6 times larger than the maximum CA IWP. Interestingly, CA IWP reaches its maximum ~ 1 h before CAC. More importantly, the maximum UTH occurs 5 h after peak convection, but within an hour of the occurrence of the maximum CAC and CA IWP values. This short time lag is the first indication that cirrus sublimation cannot be a significant source of UTH. The interpretation of these mean curves, however, is complicated by the fact

that they combine DC events of different magnitudes. Therefore, it is instructional to stratify the data according to convective strength.

d. Stratification by convective strength

Convective events are binned according to either their maximum deep convective cloud fraction $DCC_{\max}$ or maximum deep convective cloud area $DCA_{\max}$, and then averaged. The resulting Lagrangian evolution of CA IWP and UTH in the fixed target box is given in Figs. 5a,b, respectively, while that of the CA area obtained by the region-growing technique is shown in Fig. 5c. As convection gets stronger both the associated CA IWP and UTH increase significantly, although the artifact resulting from the fixed box size is clearly visible at $t = 0$ for the largest $DCC_{\max}$ bins in Figs. 5a,b. However, the region-growing analysis in Fig. 5c also shows a similarly strong increase in anvil area with deep con-

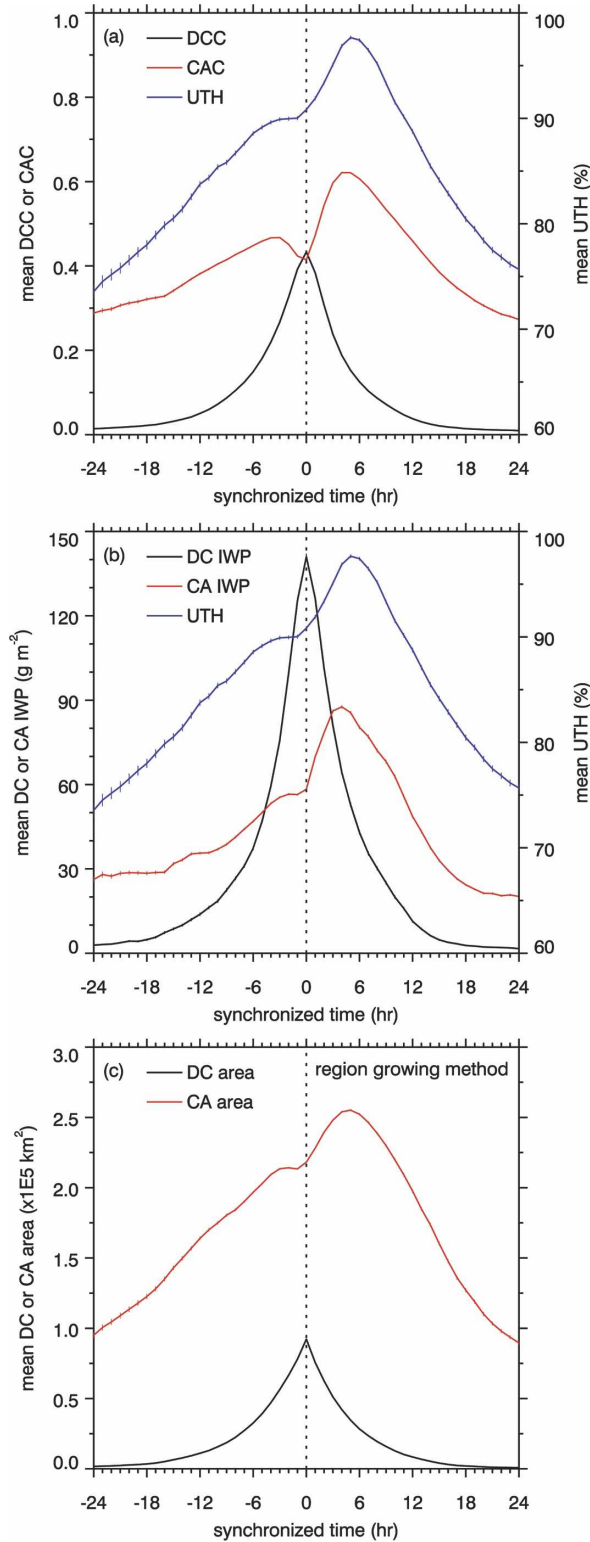


FIG. 4. Mean Lagrangian evolution of (a) DCC, CAC, and UTH; (b) DC IWP, CA IWP, and UTH; and (c) DC and CA area, averaged from all trajectories and DC events. Here, (a) and (b) correspond to a fixed-target box, while (c) shows results for the region-growing algorithm. Vertical bars depict the standard errors of the means. Peak convective activity occurs at $t = 0$.

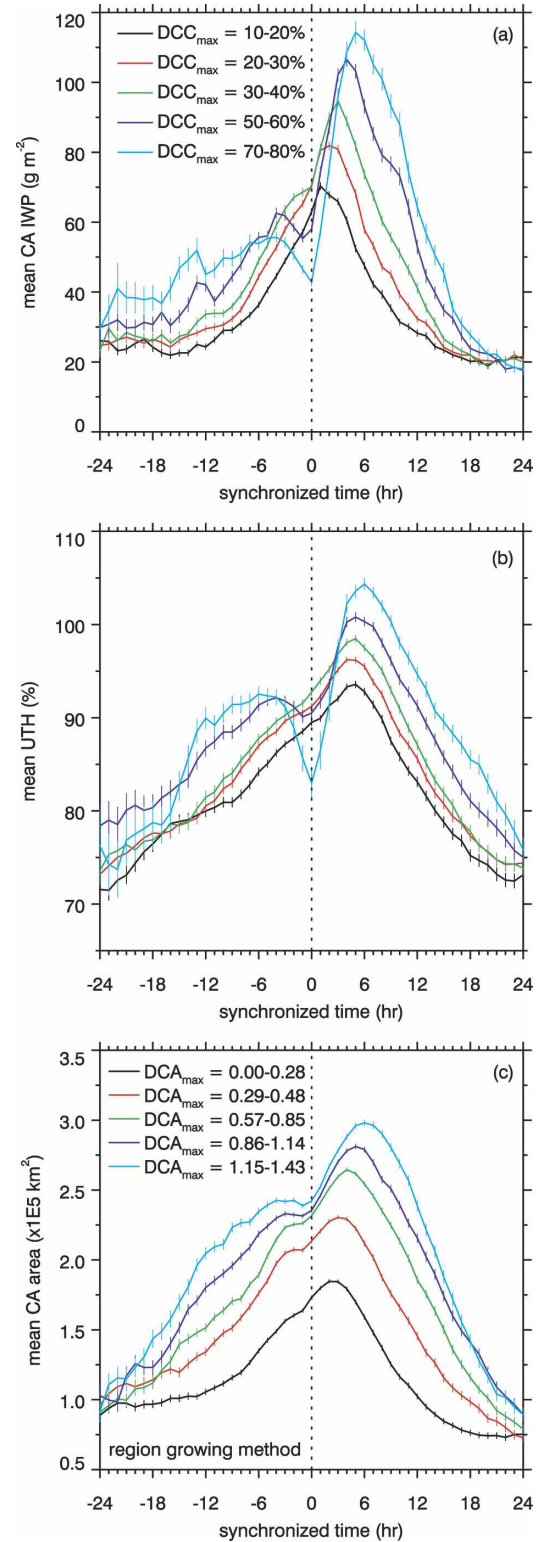


FIG. 5. Mean Lagrangian evolution of (a) CA IWP and (b) UTH binned according to maximum convective fraction $DCC_{\max}$, and (c) CA area binned as a function of maximum convective area $DCA_{\max}$. Here (a) and (b) correspond to a fixed target box, while (c) shows results for the region-growing algorithm.

vective area confirming the robustness of the fixed-box results.

Importantly, there is a marked difference between the time evolution of CA area and IWP and that of UTH. Anvil area and IWP peak systematically later for larger convective clouds, with the spreading time increasing from 1–2 h for the weakest to 5–6 h for the strongest systems. The lag in UTH, on the other hand, does not show a similar dependence on convective strength because it is always 4–5 h. Consequently, the time difference between the peak values of CA IWP and UTH is typically no more than 1–2 h.

If anvil ice were an important source of UTH, one would expect the time delay in the UTH peak to show a similar dependence on the size of convection. In addition, the lag between the UTH and CA IWP curves should also be large enough to allow the sublimation of considerable amounts of ice condensate. The results above indicate the contrary. A caveat: the analysis thus far has been based on the mean evolution of quantities. However, calculating time constants from averaged fields does not generally yield the correct average values of those time constants. Therefore, it is a more meaningful exercise to determine the anvil-spreading times, e -folding times, and lifetimes for each DC event separately and then perform the averaging. This analysis is given in the next section.

e. Average anvil-spreading and e -folding times

Spreading time scales with respect to cirrus cloud fraction, cirrus ice amount, and humidity have been defined as the time difference between DCC_{max} and CAC_{max} , DCC_{max} and $CA\ IWP_{max}$, and DCC_{max} and UTH_{max} , respectively (see Fig. 1b). All of these quantities strongly depend on convective strength as demonstrated by Fig. 6a, showing the average spreading times for CAC, CA IWP, and UTH. There is no indication of significant biases in these fixed-box results as confirmed by Fig. 6b, which plots CA area spreading times from the region-growing method. Both plots reveal that stronger storms exhibit an increasingly larger lag between peak convection and peak anvil extent/amount with the mean anvil-spreading time scale doubling from the weakest to the most intense storms.

In addition, the maximum CAC increasingly lags behind the maximum CA IWP as the strength of convection increases (see plus signs in Fig. 6a). For the smallest cloud systems, the maximum extent of the cirrus anvil roughly coincides with the maximum amount of ice condensate. For the most vigorous convection, however, the maximum ice water content precedes the peak cirrus cloud cover by approximately 2 h. Also note that the average time delay in UTH_{max} remains constant at

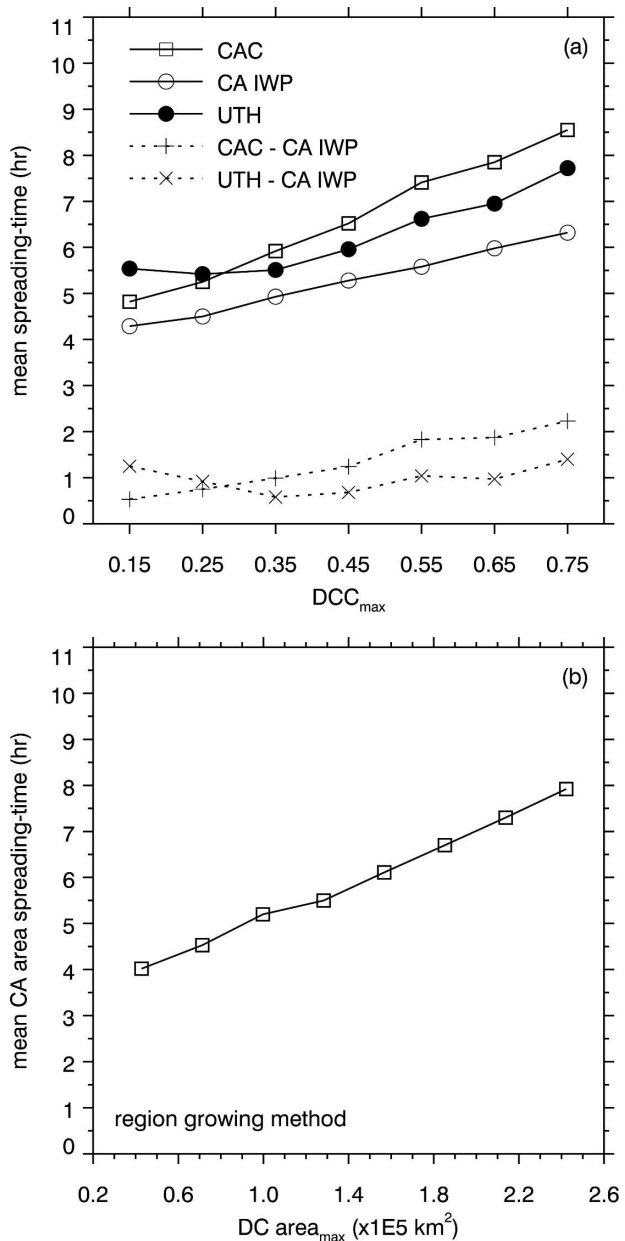


FIG. 6. (a) Mean anvil-spreading time with respect to CAC (squares), and CA IWP (circles) and mean UTH spreading time (filled circles) as a function of maximum DC cloud fraction in the fixed-target box. The difference between CAC and CA IWP spreading times (pluses) and UTH and CA IWP spreading times (crosses) are also plotted. (b) Mean CA area spreading time vs maximum DC area calculated by the region-growing method. Note that these mean values were computed from all individual DC events and not from the mean curves in Fig. 5.

5–6 h for $DCC_{max} < 0.5$, but increases for larger storms. This behavior is masked in the mean UTH curves (Fig. 5b), which all show a time delay of 4–5 h, irrespective of DCC_{max} . Finally, and most importantly, the humidity

maximum occurs typically within 1.0 h of the maximum ice water content, regardless of convective strength (see crosses in Fig. 6a). This time interval seems too small to allow for significant sublimative moistening. However, what exactly is the temporal scale of cirrus dissipation in these cloud systems?

To answer this question, e -folding decay and growth times are computed for each DC event. The e -folding decay time is the interval in which approximately two-thirds of the ice condensate dissipates from the upper troposphere via either sublimation or precipitation. The mean e -folding times for DCC, CAC, and CA IWP are shown in Fig. 7a versus the maximum DC cloud fraction in the fixed target box ($DCC_{\max}$). All variables exhibit a slightly skewed evolution about their respective peaks with decay being somewhat (0.5–1.0 h) faster than growth. The DCC e -folding times steadily increase with the size of convection, that is, larger cloud systems grow and decay more slowly and have a longer lifetime than smaller ones. [A similar behavior was previously observed by Machado et al. (1998).] When binned as a function of increasing $DCC_{\max}$, CAC e -folding times first increase, and then level off or even decrease, while CA IWP e -folding times show little variation with values around 4–5 h. However, when CAC and CA IWP e -folding times are binned according to the maximum CA cloud fraction ($CAC_{\max}$), they also show a general tendency to increase with size (Fig. 7b). The robustness of the above fixed-box results is confirmed by the region-growing method in Fig. 7c, which plots DC and CA area e -folding times versus maximum DC area. Although these e -folding times are slightly longer, they show the same general tendencies as Fig. 7a.

In conclusion, the e -folding times and, thus, lifetime of CAC tend to be significantly longer than those of DCC and CA IWP, especially for the largest cloud systems. Consequently, cirrus anvil can cover considerable areas even hours after most of its ice condensate and its parent deep convective core have dissipated. Furthermore, the ~ 4 -h mean decay time of cirrus ice is too long compared with the typically no more than 1-h average time difference between $UTH_{\max}$ and $CA\ IWP_{\max}$ to allow for significant sublimative moistening.

f. Humidity tendencies

To strengthen our argument that sublimation of cirrus condensate does *not* play a major role in directly controlling the vapor budget in the 500–200-mb layer, mean UTH tendencies are presented as a function of CAC and CA IWP tendencies (see Figs. 8a,b, respectively). Here, the Lagrangian humidity tendency is expressed as a fractional change, that is, $\Delta \ln UTH = \ln UTH(\tau + \delta t) - \ln UTH(\tau)$, while CAC and CA IWP

tendencies are given in absolute values. Calculations have been made for a time increment of $\delta t = 1, 4$, and 7 h, with the latter two values representing characteristic fixed-box decay times for CA IWP and CAC, respectively.

As shown in Fig. 8a, the average UTH tendency is positively correlated with the average CAC tendency; thus, expanding (contracting) cirrus anvil is indicative of a moistening (drying) upper troposphere. Figure 8b indicates that for short time intervals, the humidity tendency is largely independent of CA IWP tendency, and oscillates around zero. However, as δt is increased to better represent the decay time scale of cirrus condensate in the target box, the dependence of UTH tendency on CA IWP tendency becomes qualitatively similar to its dependence on CAC tendency. In both cases, the drying and moistening rates increase with the length of the time increment. Note that UTH tendencies are much larger vis-à-vis CAC tendencies than vis-à-vis CA IWP tendencies, underscoring that change in anvil cover, rather than ice content, is the better proxy for local upper-tropospheric drying/moistening.

In sum, the tendency analysis reveals no evidence for significant sublimative moistening on the time and spatial scales (1–7 h and 255 km, respectively) considered in this study. In fact, it is found that cirrus dissipation is accompanied by a drying of the upper troposphere, a result opposite that expected if cirrus ice were a primary water vapor source. Our findings suggest instead that the mean vapor budget on GCM gridbox scales is mostly influenced by dynamical mechanisms that govern the formation and decay of cirrus anvils. We note, however, that sublimation might still be important locally at cloud edges or on longer time scales.

g. The effect of sea surface temperature

The results presented so far include all DC events over both land and ocean. This section considers marine areas only, investigating what effect, if any, the underlying sea surface temperature has on the Lagrangian development of deep convective systems. For each DC event a *triggering SST* is defined as the tracking-box-mean sea surface temperature at peak convection. To get meaningful statistics trajectories are then grouped together in the following three SST bins: cold (lower quartile), medium (quartile between 37.5% and 62.5%), and warm (upper quartile), each with a sample size of ~ 6800 trajectories. The increase in mean SST from the cold to the warm bin is $\sim 2^\circ\text{C}$, with the medium bin being halfway in between. Although a 2°C dynamic range may not seem large, one should keep in mind that it is comparable to the predicted global mean surface temperature rise due to a doubling CO_2 .

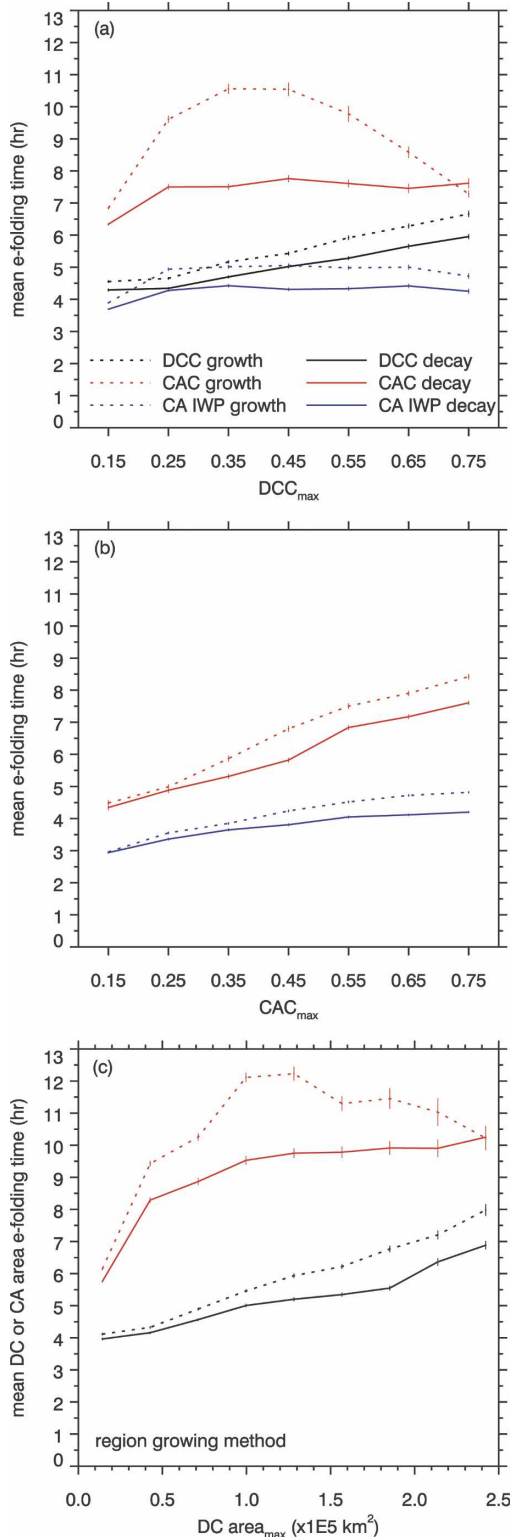


FIG. 7. Mean e -folding times vs (a) maximum DC cloud fraction and (b) maximum CA cloud fraction in the fixed-target box, and vs (c) maximum DC area calculated by the region-growing method. Dotted lines correspond to growth and solid lines refer to decay, with DCC (or DC area; black), CAC (or CA area; red), and CA IWP (blue) represented.

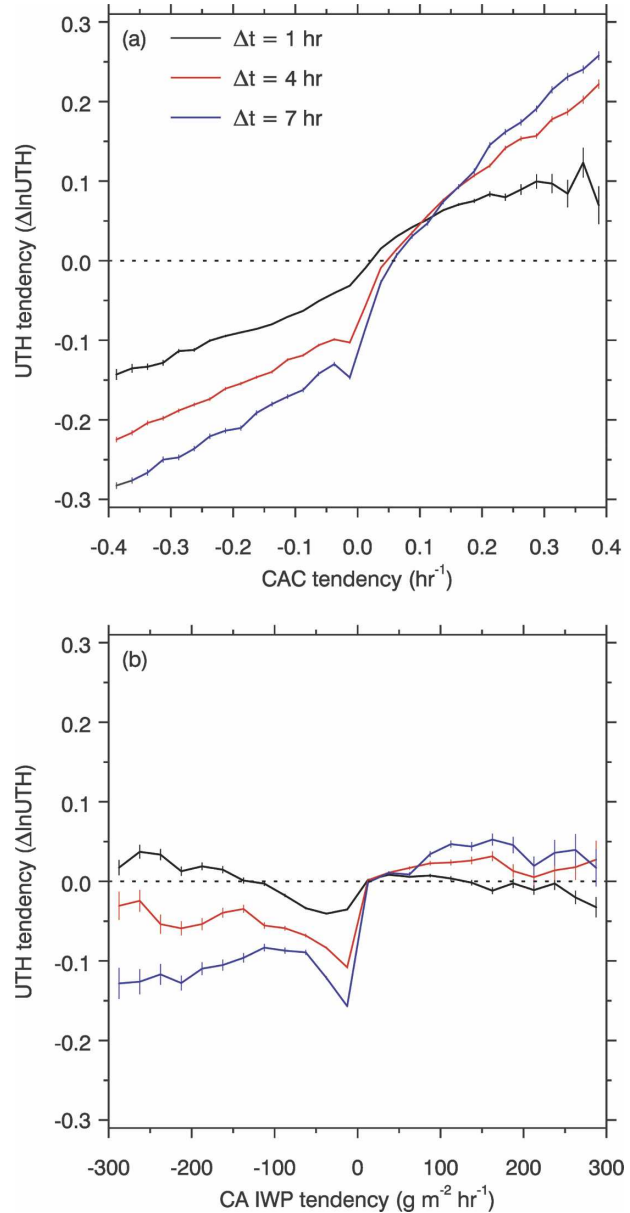


FIG. 8. Lagrangian UTH tendency binned as a function of the corresponding Lagrangian tendency in (a) cirrus anvil fraction and (b) cirrus anvil ice water path in the fixed-target box. Results are shown for a time interval of 1, 4 (characteristic CA IWP decay time), and 7 h (typical CAC decay time).

The monthly total number of tracked DC targets as a function of monthly mean SST is plotted in Fig. 9a for $3^\circ \times 3^\circ$ latitude–longitude boxes. As shown, deep convection does not occur below a monthly mean SST of 27.5°C , consistent with previous studies (Graham and Barnett 1987); however, its frequency increases sharply with SST above this threshold, resulting in a correlation coefficient of 0.77. Region-growing calculations plotted in Fig. 9b indicate that DC clouds become not only

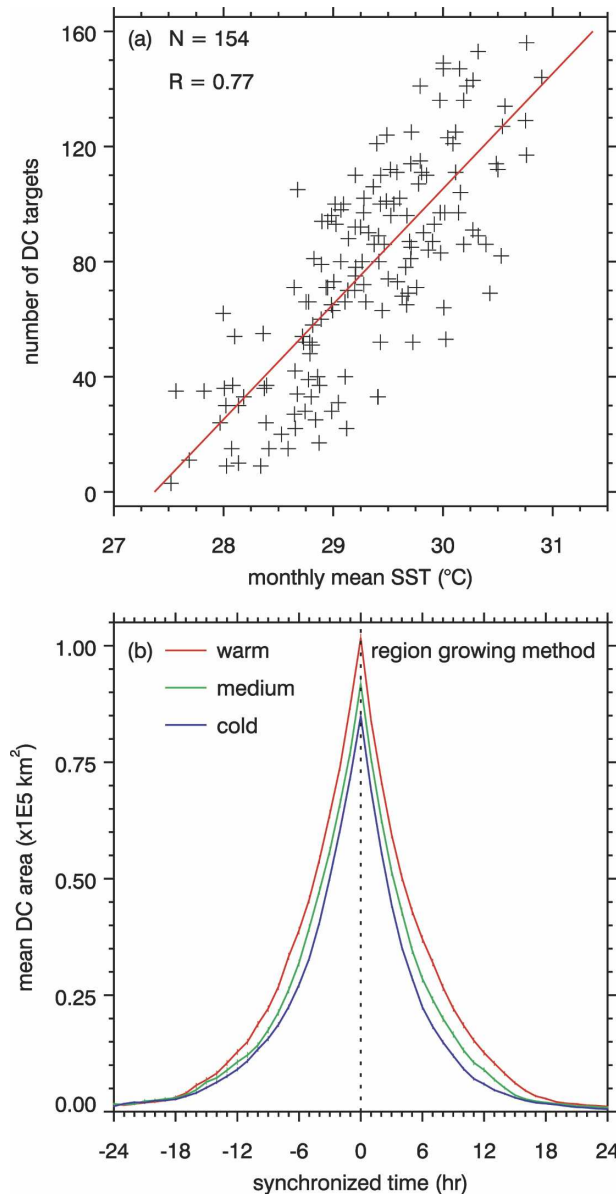


FIG. 9. (a) Total number of tracked DC targets in $3^\circ \times 3^\circ$ latitude–longitude bins as a function of mean SST for January 2006. The red line is a linear fit to the data. (b) Mean Lagrangian evolution of DC area as calculated by the region-growing method for three SST bins.

more frequent but also considerably larger over warmer oceans. In fact, the maximum DC area increases by almost $\sim 20\%$ between the cold and warm SST categories. The maximum CA area also increases with SST although only slightly, as depicted in Fig. 10a. However, there is a more pronounced increase in CA coverage over warmer oceans hours after the CA peak as a result of longer anvil-spreading times and slower decay times. Finally, Fig. 10b plots the evolution of

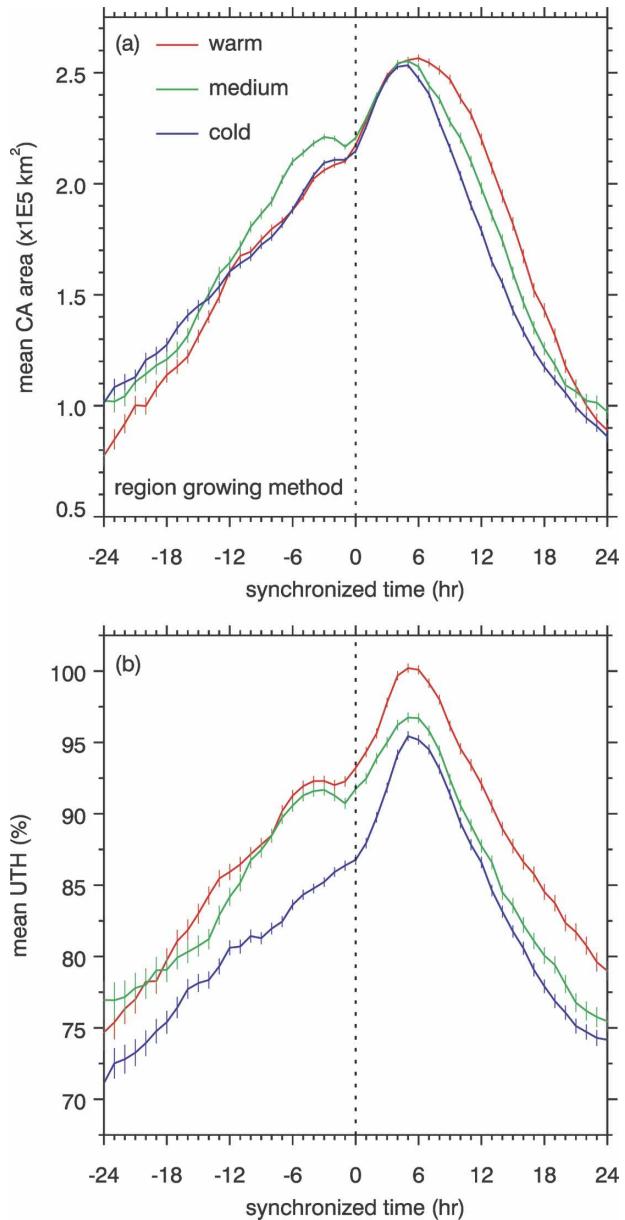


FIG. 10. Mean Lagrangian evolution of (a) CA area and (b) UTH for three SST bins. Note that UTH is calculated in the fixed-target box, while CA area is computed with the region-growing method.

UTH, which shows the most sensitivity to SST. Storms developing over warmer oceans are clearly associated with larger UTH values.

The analysis presented thus far has considered absolute values of CA area and UTH averaged over DC events of all different sizes. As deep convection becomes more extensive with SST, this complicates the interpretation of results. Therefore, it is worthwhile to isolate the changes in CA area and UTH evolution that

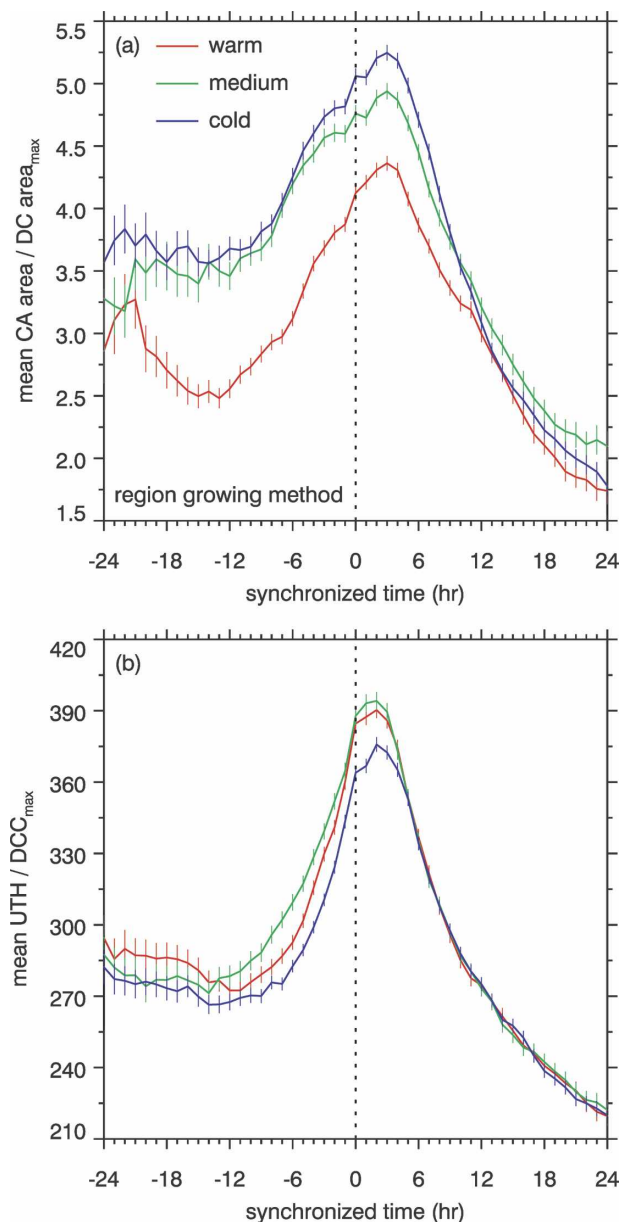


FIG. 11. Same as Fig. 10, but normalized by the strength of convection $DC_{area_{max}}$ or DCC_{max} .

are primarily due to SST effects, from those due to the varying magnitude of convection. This can be achieved through normalizing these quantities by the maximum DC extent, as given in Fig. 11 for the three SST classes. (The plotted parameters are cirrus anvil area and UTH per unit cumulonimbus area.) Although the areas of both deep convective core and cirrus anvil increase with SST, the efficiency at which convection generates cirrus anvil appears to decline as SST increases (Fig. 11a).

Given the fact that our high cloud classification is based on a fixed brightness temperature threshold

separating deep convection from anvil cirrus, it is possible that the apparent reduction in anvil detrainment efficiency is a consequence of cloud misclassification resulting from a decrease in cloud-top brightness temperature with SST. As discussed in section 3b, cloud tops do become somewhat colder over warmer oceans; however, the threshold brightness temperature dividing high cloud clusters remains quite constant at approximately $T_{11} = 215$ K and the mean anvil-top height stays at ~ 11 km (~ 250 mb) for all three SST bins. Furthermore, the SST dependence of cirrus detrainment efficiency shows the same tendency for a reasonable range of threshold brightness temperatures between 215 and 220 K, building further confidence in the above results.

In contrast, the mean evolution of normalized UTH shows essentially the opposite sensitivity to SST, as depicted by Fig. 11b. Its peak value tends to be larger for warmer oceans, suggesting that while the efficiency at which cirrus anvil is detrained from a convective event decreases as SST increases, it has little impact on UTH. This finding is consistent with our earlier conclusion that UTH is determined primarily by dynamical transport from convective updrafts, rather than by sublimation of detrained anvil ice water.

5. Summary and discussion

In this paper the complex relationship among upper-tropospheric humidity, tropical deep convection, and sea surface temperature has been investigated with specific emphasis on the following two debated issues: (i) the importance of cirrus anvil sublimation in moistening the upper troposphere, and (ii) the effect of SST on the development of deep convective systems. The study has been performed from a Lagrangian perspective, whereby the coevolution of water vapor in the 500–200-mb layer and cloud macro- and microphysical properties has been analyzed as a function of the convective life cycle using geostationary satellite observations. A large number of storms have been tracked over the tropical western Pacific in January 2006 in order to obtain robust statistics about their average development. Artifacts and potential biases resulting from using a fixed box size in the analysis have been evaluated with the help of region-growing calculations, confirming all major conclusions drawn from gridded results.

A traditional Eulerian analysis has revealed a strong relationship between monthly mean UTH and cirrus ice water path with correlations in the range of 0.7–0.8. Similarly strong positive correlations between these quantities have already been observed by previous investigators and successfully reproduced by GCMs, seemingly lending observational evidence to predic-

tions by one-dimensional models of tropical convection, where the distribution of upper-tropospheric moisture is very sensitive to the sublimation of cloud condensate. The Lagrangian analysis, however, has painted a different picture. The mean evolution averaged from all deep convective events has indicated that UTH reaches its maximum ~ 5 h after peak convection, while cirrus anvil cover and ice water amount peak at ~ 0.5 h and ~ 1 h prior to that, respectively. The short time lag between the mean UTH and ice water path development provides the first indication that cirrus condensate cannot be the source of the elevated humidity.

A further stratification of the results has shown that the time lag between peak convection and both peak anvil area and cirrus ice amount steadily increases with storm size. Simultaneously, the time difference between maximum anvil area and maximum cirrus ice amount also increases, with the latter occurring first. However, UTH is tightly synchronized with cirrus ice water path lagging behind it by no more than 1.0 h, irrespective of the strength of convection. Considering that the characteristic decay time of cirrus ice condensate has been found to be ~ 4 h, this short time interval cannot allow for significant sublimative moistening.

A further piece of evidence has been obtained by analyzing UTH tendencies against tendencies in cirrus anvil cover and cirrus ice amount. A strong positive correlation has been found between UTH and cirrus cover tendencies, confirming earlier results by Soden (2004). The UTH tendencies, however, have been generally much smaller when binned against ice water path tendencies, and either have oscillated around zero or have been positively correlated with ice water path tendencies, depending on the length of the tendency time increment. Contrary to what one would expect if cirrus ice were a primary water vapor source, these results have shown that cirrus decay (in terms of both cloud cover and integrated ice content) is accompanied by a drying of the upper troposphere, while cirrus growth is indicative of increasing UTH.

Taken together, the above findings add to the growing body of evidence strongly suggesting that sublimation of cirrus anvil condensate does *not* play a significant role in modulating the vapor budget of the 500–200-mb upper-tropospheric layer. The elevated humidity levels observed a few hours after peak convection seem instead linked to vertical transport processes responsible for convective detrainment of water vapor and the formation and maintenance of cirrus anvils.

The effect of sea surface temperature on Lagrangian cloud evolution has also been investigated with the help

of newly available high spatial and temporal resolution microwave SSTs. Deep convective trajectories have been sorted into cold, medium, and warm categories based on their triggering SSTs, with a $\sim 2^\circ\text{C}$ mean temperature difference between the two extreme bins. It has been found that a rise in SST of such magnitude results in a measurable increase in the frequency, spatial extent, and water content of deep convective cores. These larger storms over warmer oceans are associated with cirrus anvils whose maximum extent is also slightly larger and whose decay time is somewhat slower than their counterparts over colder oceans. However, normalization by the strength of convection has shown that anvil area per unit cumulus area, that is, cirrus detrainment efficiency, actually decreases as SST increases. Finally, both the mean UTH and its detrainment rate have generally increased with SST.

It is noted that a qualitatively similar sea surface temperature dependence of cirrus anvil detrainment efficiency has been observed by Lindzen et al. (2001). They have found that, when large areas of the tropics are considered, there is a strong inverse relation between normalized anvil cloud fraction and the underlying mean SST, and they interpreted the result as reduced cirrus outflow caused by an increase in deep convective precipitation efficiency over warmer oceans. Combining this area effect with a particular set of assumptions on the mean radiative properties of high-level clouds, Lindzen et al. (2001) then proposed a strong negative “iris” feedback that may operate to minimize global warming due to a doubling of CO_2 . However, all aspects of the iris mechanism have been questioned in the literature. For example, Lin et al. (2002) have found significantly larger cloud albedos and longwave fluxes in state-of-the-art satellite measurements than those assumed by Lindzen et al. (2001), resulting in a weak positive feedback instead of a strong negative feedback. Furthermore, Hartmann and Michelsen (2002) have pointed out that the observed relationship between cloud-weighted SST and normalized anvil fraction results from changes in cloud amount over colder SSTs, which is far removed from tropical deep convection whose anvil clouds Lindzen et al. (2001) have hypothesized are modulated by small SST variations.

Our study does not endorse the iris feedback; it simply notes an apparent decrease in anvil detrainment efficiency with SST in our analysis. Unfortunately, the employed infrared data are not suitable for studying changes in convective precipitation efficiency, which have been proposed as the cause of the area effect. However, one can still point out that while Hartmann and Michelsen’s (2002) explanation is plausible when cloud fractions and cloud-weighted SSTs are calculated

for large geographical areas, it cannot explain the present results obtained by tracking individual convective events of relatively small sizes.

The most recent investigations of the area effect remain divided. On the one hand, the study by Rapp et al. (2005) combining TRMM precipitation radar data with infrared imagery has found no significant correlations between SST and deep convective cloud size normalized by rainfall amount, providing further evidence against iris. On the other hand, the multisatellite study of Spencer et al. (2007) is supportive of iris and suggests an increase in rainfall efficiency as the troposphere warms.

Our findings also indicate an area effect. Although we have eliminated gridding (fixed box) artifacts, we cannot rule out the possibility that this result is due to the combination of a fixed-temperature threshold and higher-colder cloud tops over warmer oceans. In fact, a recent paper by Kubar et al. (2007) analyzing coincident optical cloud retrievals and microwave precipitation rates seems to provide some indirect evidence to that effect. They have found nearly identical deep convective cloud amounts, but more anvil clouds per unit of precipitation in the warmer west Pacific than in the colder east Pacific, suggesting an anti-iris effect. It remains to be seen, however, if the same SST dependence holds true when the analysis is restricted to the west Pacific, as in our case. In addition, the Kubar et al. (2007) study is limited by a cloud classification also employing fixed thresholds (on optical thickness) and microwave retrievals that assume a specific analytical relationship between rain rate and cloud water amount/optical thickness (Wentz and Meissner 2000).

Ultimately, all passive imagery-based cloud classifications suffer from errors introduced by thresholding continuous distributions of cloud-top parameters. Active sensors can provide cloud vertical cross sections and might be better suited to identify the anvil and convective parts of tropical cloud systems. Therefore, further studies using the new A-Train dataset (Stephens et al. 2002) are needed for the definitive treatment of the subject.

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